FAILURE STRENGTH OF ICY LITHOSPHERES

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It has been widely assumed in the literature that the stresses required for brittle failure of an icy lithosphere under tension are limited by the tensile strength of intact ice. This assumption has led to the widespread use of 2 MPa (20 bars) for the extensional strength of predominantly ice lithospheres (e.g. Ganymede's), based on the unconfined tensile strength of ice near its melting temperature measured by Hawkes and Mellor (1). However, our understanding of the maximum stress levels in the earth's lithosphere are based on the frictional resistance to sliding on pre-existing fractures and not the strength of intact rock in laboratory experiments, which is always greater. Similar friction relations for ice predict maximum stresses that are substantially greater than the apparent tensile strength of intact ice near its melting temperature. At first inspection, then, it appears that ice rich lithospheres would fail under tensional stress by tensile fracture of intact ice rather than sliding on pre-existing fractures. In this abstract we will introduce lithospheric strengths derived from friction on pre-existing fractures and ductile flow laws, derive these relations for icy lithospheres, show that the the tensile strength of intact ice under applicable conditions is actually an order of magnitude stronger than widely assumed, and demonstrate that this strength is everywhere greater than that required to initiate frictional sliding on pre-existing fractures and faults.

The maximum stress levels found in the earth's crust are accurately predicted by Byerlee's law (2,3). This relation is based on laboratory measurements of the frictional resistance to sliding on pre-existing fractures, which occurs at stresses less than those required to break intact rock. Byerlee's law is of the form $t=uS_n+b$, where t is the shear stress, S_n is the effective normal stress (normal stress minus pore pressure), u is the coefficient of friction, and b is a constant. In terms of S_1 and S_3, the maximum and minimum principal effective stresses (stress minus pore pressure), Byerlee's law can be written $S_1=KS_3+B$, where $K=[(u^2+1)^{1/2}+u]^2$, B is a function of b and u, and sliding is assumed to occur on the most advantageous slip plane. Laboratory friction measurements show that u and b are virtually independent of stress (except for a slight change for rocks at 135 MPa), rock type, displacement, surface conditions, and temperature. For this application we will use the friction law for low stress determined by Byerlee (2) $t=0.85S_n$, which has been found to hold for a wide variety of geologic materials, because the appropriate low normal stress friction measurements on ice have not been made. At higher stress we use the friction data measured for ice, u=0.2 and b=10 MPa (4). The higher stress friction data for ice do constrain the applicable low stress friction law to be very close to that for rock, because the lowest normal stress measurements for ice are for S_n=17 MPa (5), which constrains the low stress friction law, that also must pass through the origin, to have a slope only marginally different ($u \geq 0.79$) from that determined for rocks (u=0.85). In order for fluid pore pressure to decrease the friction on pre-existing faults and fractures, thereby changing the slope of the brittle yield stress versus depth curve, liquid water must fill the connected pore space. This is not likely to have occurred on the icy satellites given the extremely low surface temperatures (less than 100°K) and reasonable thermal gradients. Because the vertical stress is generally quite close to the lithostatic load (density times gravity times depth) in areas of low relief, this relation

predicts a linear increase in yield stress with depth (Fig. 1).

With increasing temperature, rock and ice deformation occurs by ductile flow. Flow laws for rocks, minerals, and ice have been experimentally determined for stresses up to 1-2 GPa and strain rates down to 10^{-8}/sec. These results can be extrapolated to geologic strain rates via creep equations which generally are of the form $de/dt=A(S_1-S_3)^n\exp(-Q/RT)$, where de/dt is the strain rate, R is the gas constant, T is absolute temperature, and A, Q (the activation energy), and n are experimentally determined constants. As a result the ductile strength is negligible at depths where T is high and increases exponentially with decreasing depth. We have used the experimentally determined flow parameters (6,5) for pure ice I_h, $A=1.2\times10^{-24}$/sec-Pa^4, $Q=4.5\times10^4$ J/mole and $n=4.0$ extrapolated to geologic strain rates of 10^{-15}/sec (about 3%/m.y.). A surface temperature of 100°K and a thermal gradient of 1.6°/km applicable to the early high temperatures likely in some of the larger icy satellites (e.g. Ganymede, 7) yields the strength envelope illustrated in Fig. 1. It has been shown at high temperatures (i.e. polar conditions on the earth) that the inclusion of a small amount of silicates will greatly increase the creep strength of ice I_h (8). The amount of hardening from the inclusion of small amounts of silicates in the lithospheres of the icy satellites is not known, but for lack of a better constraint we will assume that the hardening resulting from the addition of less than a few percent of silicates in icy satellites lithospheres can be bracketed by an order of magnitude increase in creep strength (e.g. 9).

The failure criterion for a given depth in the lithosphere is determined by the weaker of the frictional or ductile strength at that depth. The yield stress increases with depth according to Byerlee's law until it exceeds that calculated using the appropriate flow law, after which it decreases exponentially with depth. The intersection of the brittle and ductile yield stress curves defines the brittle-ductile transition depth and also the peak stress needed to cause failure of the entire lithosphere. As a result the peak stress needed for lithospheric failure is dependent on the thermal gradient, which causes the ductile flow law to intersect the friction curve at different depths (shallow for high thermal gradients, deep for low thermal gradients). These peak stresses are probably somewhat greater than those actually required to cause failure of the lithosphere because semibrittle and low temperature ductile processes tend to round off the intersection points between the brittle and ductile curves (10).

As can be seen in Fig. 1, the peak stress needed to cause tensile failure of the lithosphere using the above parameters is on the order of 10 MPa (100 bars), although the average stress in the lithosphere is only about half that value. This strength is applicable for large icy satellites (e.g. Ganymede). We have also determined strengths for the smaller icy satellites of Saturn using the appropriate gravities, densities, surface temperatures, and calculated thermal profiles (11) and found all strengths are substantially larger than this value. As a result this value is probably a minimum for the lithospheres of the icy satellites.

Note that this peak stress (10 MPa) is significantly higher than the tensile strength of about 2 MPa used in virtually all previous studies, which is based on the the unconfined tensile fracture strength of ice near its melting temperature. Because failure should occur by the mechanism requiring the lowest stress, it would appear on first inspection that tensile fracture of intact ice is the relevant mode of fracture for icy satellites. However it can be shown using a Griffith failure criterion and taking the vertical lithostatic load into account that the stress difference for failure due to the tensile fracture of intact material at depth z is given

by $S_1-S_3=S_o+pgz$ for $S_1-S_3<4S_o$ and by $S_1-S_3=4[(S_o pgz+S_o^2)^{1/2}-S_o]$ for $S_1-S_3>4S_o$, where p is density, g is gravitational acceleration, and S_o is the unconfined tensile strength (e.g. 12). The first of these relations describes the opening of tension cracks, which occurs when the confining pressure is relatively low. The second relation describes shear failure in tension (or compression), which is the mode of failure when the confining pressure is too great for open tension cracks to form. The depth of transition between the two modes of tensile failure is given by $z=3S_o/pg$ (about 5 km for S_o=2.5 MPa on a satellite with a gravitational acceleration similar to that of Ganymede). The curve for failure due to fracture of intact ice is plotted in Fig. 1, where it can be seen that frictional failure is preferred at all depths for the assumed material parameters of ice. This conclusion is borne out by the experimental results, in which ice samples failed due to frictional sliding along pre-existing saw cuts rather than initiating new fractures (4). Note also that if the tensile fracture strength of intact ice increases with decreasing temperature, as is the case for the compressional fracture strength (13), the curve for fracture of intact ice under tension will move to the left and frictional failure will be even more favorable. Similar results are also obtained for other choices of failure criteria. Thus the failure strength of an icy lithosphere is significantly greater than has been previously assumed.

In conclusion, because the tensile strength of intact ice increases markedly with confining pressure, it actually exceeds the frictional strength at all depths. Thus, icy lithospheres will fail by frictional slip along pre-existing fractures at yield stresses greater than previously assumed rather than opening tensile cracks in intact ice.

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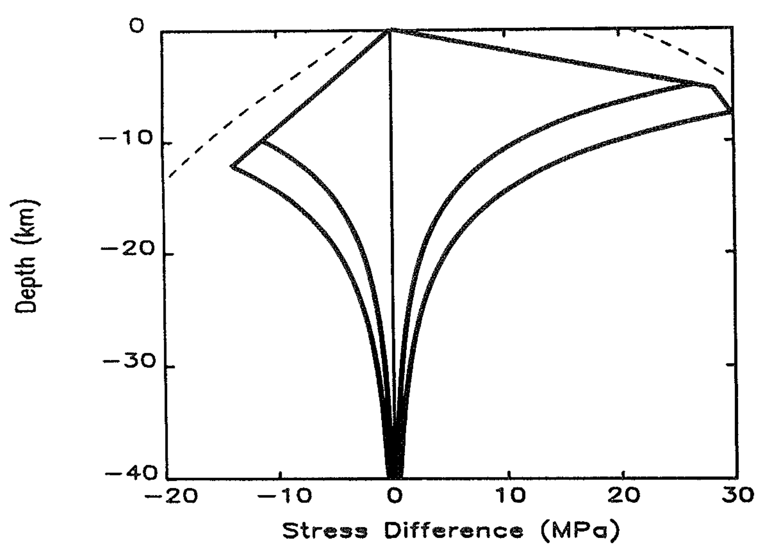

Fig. 1. Brittle and ductile yield stress versus depth curve for compression (to the right) and extension (to the left), with a peak stress of 10 MPa (see text for discussion). The dashed lines show the failure strength of intact ice as a function of depth assuming a Griffith failure criterion. Because stresses required to break intact rock are greater than those required to initiate sliding on pre-existing fractures, lithospheric failure occurs by frictional sliding on pre-existing fractures.